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Laurentian Great Lakes Dynamics, Climate, and Response to Change

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1. Introduction

The Laurentian Great Lakes are one of North America's greatest water resources. Due to recent record high lake levels and to climate change issues, there is renewed interest in lake level trends and in factors affecting high water levels. Impacts on Great Lakes water supply components and basin storages of water and heat must be understood before lake level impacts can be assessed. The Great Lakes Environmental Research Laboratory (GLERL) developed conceptual simulation models for Great Lakes hydrology to address the impact questions. GLERL integrated the models to estimate lake levels, whole-lake heat storage, whole-basin moisture storage, and water and energy balances for forecasts and for assessment of impacts associated with climate change. Because the Great Lakes possess tremendous water and heat storage capacities, they respond slowly to changed meteorologic inputs. This "memory" results in a filtering or dampening of most short-term meteorologic fluctuations and in a response to longer-period fluctuations characteristic of climate change. The large Great Lakes system, thus, is ideal for studying regional effects of climate changes.

This paper outlines large lake dynamics and climate, pertinent especially to the Laurentian Great Lakes, summarizes GLERL's Great Lakes hydrology models, presents results of recent climate change studies on the Great Lakes, and looks ahead to the next generation of interactively coupled models useful for assessing climate change.

2. Great Lakes Dynamics and Climate

There is a major tendency to think of Great Lakes water levels in terms of extremes rather than of normal conditions. Within recent memory we had the record low lake levels of 1964. This resulted in docks sitting out of the water, insufficient depths for navigation in many harbors and channels, and greatly reduced recreational opportunities. These low levels were followed in 1973 by record high lake levels with resultant flooding and shore damage and erosion. The lake levels remained high until 1989, whence they returned to near-average

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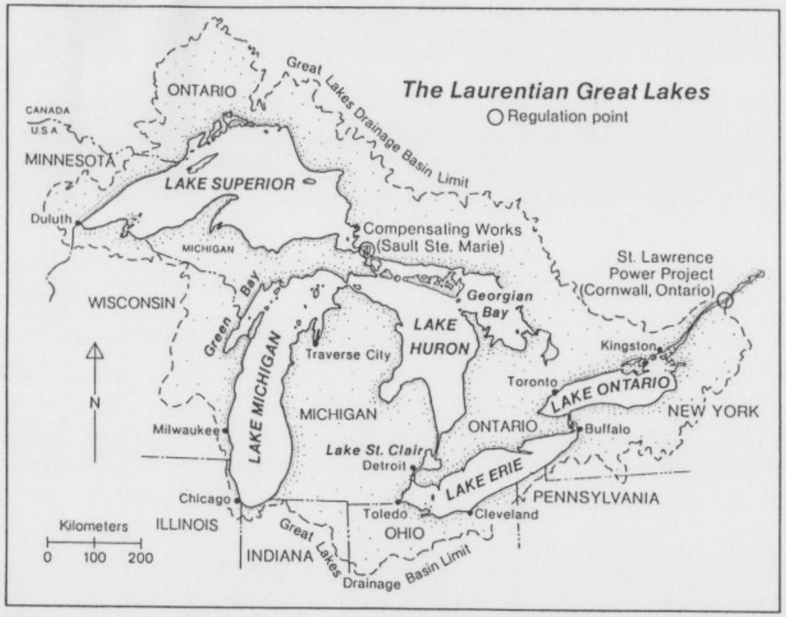


Figure 1. Great Lakes Basin.

conditions, and new record highs were once again set on Lakes Superior, Michigan-Huron, St. Clair, and Erie.

This section presents an overview of the physical characteristics of the Great Lakes from a water quantity perspective, outlines the basin and lake physical processes, summarizes the climatology of the Great Lakes, examines the types of natural lake level fluctuations and their causes, compares the natural fluctuations with existing diversions and regulation effects, describes current conditions, and concludes with a long-term perspective on lake levels.

2.1 Great Lakes Overview

The Great Lakes basin, shown in Figure 1, contains an area of approximately 770,000 km^2 (300,000 mi^2), about one-third of which is water surface. Cursory descriptions are given by *Freeman and Haras* (1978), the *U. S. Army Corps of Engineers* (1985), and the *Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data* (1977). The basin extends some 3,200 km (2,000 mi) from the western edge of Lake Superior to the Moses-Saunders Power Dam on the St. Lawrence River. The water surface drops in a cascade over this dis-

tance some 180 m (600 ft) to sea level. The most upstream, largest, and deepest lake, is Lake Superior. The lake has two interbasin diversions of water into the system from the Hudson Bay Basin: the Long Lac and Ogoki Diversions. Lake Superior waters flow through the lock and compensating works at Sault St. Marie and down the St. Marys River into Lake Huron where it is joined by water flowing from Lake Michigan. Lake Superior is completely regulated, to balance Lakes Superior, Michigan, and Huron water levels, according to Regulation Plan 1977, under the auspices of the International Joint Commission (*International Lake Superior Board of Control* 1981, 1982).

Lakes Michigan and Huron are considered to be one lake hydraulically because of their connection through the deep Straits of Mackinac. The second interbasin diversion takes place from Lake Michigan at Chicago. Here water is diverted from the Great Lakes to the Mississippi River Basin. The water flows from Lake Huron through the St. Clair River, Lake St. Clair, and Detroit River system into Lake Erie. The drop in water surface between Lakes Michigan-Huron and Lake Erie is only about 2 m (8 ft). This results in a large backwater effect between Lakes Erie, St. Clair, and Michigan-Huron; changes in Lakes St. Clair and Erie levels are transmitted upstream to Lakes Michigan and Huron. From Lake Erie the flow is through the Niagara River and Welland Diversion into Lake Ontario. The major drop over Niagara Falls precludes changes on Lake Ontario from being transmitted to the upstream lakes. The Welland Diversion is an intrabasin diversion bypassing Niagara Falls and is used for navigation and hydropower. There is also a small diversion into the New York State Barge Canal System which is ultimately discharged into Lake Ontario. Lake Ontario is completely regulated in accordance with Regulation Plan 1958D to balance damages upstream on Lake Ontario with those downstream on the St. Lawrence Seaway [estimated to have lowered Lake Ontario 0.75 m (2.5 ft) in 1986]. The outflows are controlled by the Moses-Saunders Power Dam between Massena, New York and Cornwall, Ontario. From Lake Ontario, the water flows through the St. Lawrence River to the Gulf of St. Lawrence and to the ocean.

Lakes Superior, Michigan, Huron, and Ontario are very deep, while Lakes Erie and St. Clair are very shallow. Table 1 contains pertinent gross statistics on the sizes of the Great Lakes, Lake St. Clair, and their basins.

2.2 Physical Processes

The behavior of the Laurentian Great Lakes system is governed by its huge storages of water and energy. There are three main conservation laws to consider relative to these huge storages: 1) mass balances in the basins, 2) mass balances in the lakes, and 3) energy balances in the

Table 1. Laurentian Great Lake Size Statistics.

Characteristic		Superior	Michigan	Huron	St Clair	Erie	Ontario
Basin area,	km^2	128,000	118,000	131,000	12,400	58,800	60,600
	mi^2	49,300	45,600	50,700	4,800	22,700	23,400
Surface area,	km^2	82,100	57,800	59,600	1,114	25,700	18,960
	mi^2	31,700	22,316	23,000	430	9,920	7,320
Volume,	km^3	12,100	4,920	3,540	3	484	1,640
	mi^3	2,900	1,180	850	1	116	393
Average depth,	m	147	85	59	3	19	86
	ft	482	280	190	10	62	280
Maximum depth,	m	405	281	229	6	64	244
	ft	1,330	923	750	21	210	802

lakes. There are also mass and energy balances to consider for the lakes' ice cover. The first conservation law (mass balance on the basins) comprises the primary process determining lake levels: the hydrologic cycle of the Great Lakes Basin (*Croley* 1983a). As shown in Figure 2, precipitation enters the snowpack, if present, and is then available as snow melt depending mainly on air temperature and solar radiation. Snow melt and rainfall partly infiltrate into the soil and partly run off directly to rivers, depending upon the moisture content of the soil. Infiltration is high if the soil is dry, and surface runoff is high if the soil is saturated. Soil moisture evaporates or is transpired by vegetation depending upon the types of vegetation, the season, solar radiation, air temperature, humidity, and wind speed. The remainder percolates into deeper basin storages which feed the rivers and lakes through interflows and groundwater flows. Generally, these river supplies are high if the soil and groundwater storages are large. Because of this buffering effect of the large snowpack and the large soil, groundwater, and surface storages, runoff from rivers into a lake can remain high for many months or years after high precipitation has stopped.

Mass conservation in the lake is the next major determinant of lake levels. Major sources of water into a lake include precipitation on the land basin which results in runoff into the lake, precipitation over the lake surface, inflow from upstream lakes, and diversions into the lake. Net groundwater flows directly to each of the Great Lakes are generally neglected (*DeCooke and Witherspoon* 1981). The outflows consist of evaporation from the lake surface, flow to downstream lakes, and diversions. The imbalance between the inflow and outflow results in the lake levels either rising if there is more inflow than outflow, represented by a positive change in storage, or falling if there is more outflow than inflow, represented by a

negative change in storage. The large lake water storages provide a buffering of the input fluctuations with regard to output variations. The large surface areas of the lakes enable large storage changes with very small water level changes; hence, outputs (which are a function of water levels) change slowly.

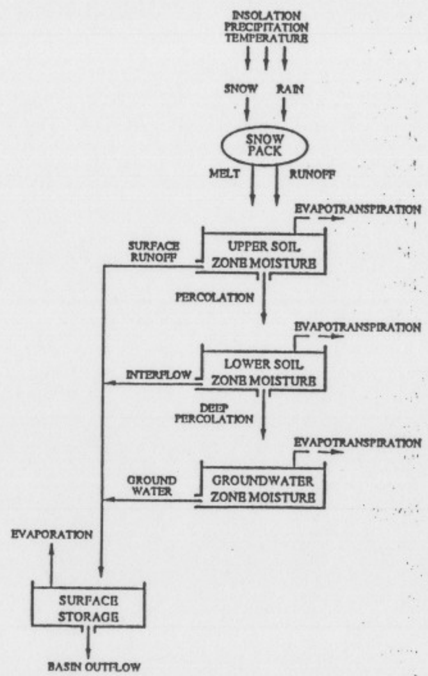


Figure 2. Runoff Hydrology Concepts.

Energy conservation in a lake actually must be considered together with a lake's mass balance. Lake heat storage is a function of the lake's size and shape and of its surface inputs of solar insolation and reflection (short wave exchanges), thermal emission and atmospheric emission (net long wave exchange), conduction to the atmosphere (sensible heat transfer), heat loss through evaporation (latent and some advection), other advection terms (precipitation, inflows, and outflows), and ice growth and melt. Evaporation is a function of surface temperature (heat storage), air temperature (atmospheric stability), humidity, and wind speed. Water surface temperatures generally peak in August (September for Superior) at 15-25 °C resulting in a stable summertime temperature stratification in the water column (high-density cool water at depth and low-density warm water at the surface). Surface temperatures drop during the fall and winter, and the water column in each lake "turns over" as temperatures drop through 4°C where water density is maximum (deep now-lighter waters rise and mix with now-heavier surface layers). Turn over occurs again in the spring as surface temperatures rise to that of maximum density.

There is also extensive ice cover on most of the lakes during most winters. Lake Superior averages about 75% ice-covered, Michigan is 45%, Huron is 68%, Erie is 45%, and Ontario is 24%. Ice formation and breakup is governed by additional mass and energy balances that take place simultaneously with those of the lakes' water bodies. The Great Lakes do not ordinarily freeze-over completely (*Assel et al.* 1983) because of the combination of their large heat storage capacity, large surface area, and their location in the mid-latitude winter storm track. Alternating periods of mild and cold air temperatures combine with episodic high and low wind stresses at the water surface to produce transitory ice conditions during the winter. Ice cover in mid-lake regions is often in motion. Lake Erie ice speeds have been observed to average 8 cm/s with a maximum speed of 46 cm/s (*Campbell et al.* 1987). Ice can form, melt, or be advected toward or from most mid-lake areas throughout the winter (*Rondy* 1976). When ice is advected into areas with existing ice cover, it can under- or override the ice cover, forming rafted rubble 5-10 m thick. The normal seasonal progression of ice formation begins in the shallow shore areas of the Great Lakes in December and January. The deeper mid-lake areas normally do not form extensive ice cover until February and March. Ice is lost over all lake areas during the last half of March and during April.

Ice formation alters the surface thermodynamics of the lakes, changing subsequent ice formation, surface heating or cooling, lake evaporation, and lake responses to atmospheric changes. The large heat storages of the lakes provide a buffering; they forestall and reduce ice formation and shift the large evaporation response. Water temperatures lag air temperatures and evaporation lags surface heating (insolation). Evaporation peaks in October-November on Lake Erie and in November-December on Lake Superior.

The large basin and lake storages of water and ice and the large lake and ice storages of energy represent an "intrinsic memory" that allow scientists to forecast basin moisture storage and runoff (basin storage buffering) in the face of uncertain meteorology. It also allows prediction of evaporation (heat storage buffering) and lake levels (lake storage buffering) of up to about six months of low-frequency changes. It further enables estimation of ice formation amounts and timing as well as all secondary hydrological variables.

2.3 *Climatology*

Precipitation causes the major long-term variations in lake levels (*Quinn and Croley* 1981; *Quinn* 1985). Table 2 shows that annual precipitation ranges from about 82 cm (32 in) for Superior to 93 cm (37 in) for Ontario. Figure 3 depicts total annual precipitation over Lakes Michigan-Huron, St. Clair, and Erie for the 1900-79 period (*Quinn* 1981; *Quinn and Norton* 1982). From 1900 through 1939, a low precipitation regime predominated with the majority of the years falling below the mean. From about 1940 until recently, a high precipitation regime has existed. Of particular interest is the high precipitation in the early 1950s, the low precipitation in the early 1960s that led to the record lows, and a consistently very high pre-

Table 2. Partial Great Lakes Annual Water Balance (1951-88).

Component	Superior		Michigan		Huron		Erie		Ontario	
	(cm)	(in)	(cm)	(in)	(cm)	(in)	(cm)	(in)	(cm)	(in)
Lake Precip.[a]	82	32	83	32	87	34	91	36	93	37
Lake Runoff[a]	62	24	64	25	84	33	80	32	169	67
Lake Evap.[a]	56	22	65	25	63	25	90	35	67	26

[a]Equivalent depth over the lake area.

cipitation regime from the late 1960s through the late 1980s. Table 3 summarizes Great Lakes annual precipitation totals by basin for several periods. Of particular interest are the progressions of increasing precipitation for each basin. While the 1940-90 period is generally above normal (2-8% higher than the 1900-69 average and -2-6% higher than the 1900-90 average), the last 20 of those years are higher still (8-13% than the 1900-69 average and 2-11% higher than the 1900-90 average); 1985 set many new records with the highest precipitation to that date (8-40% higher than the 1900-69 average and 7-33% higher than the 1900-90 average).

Variations in air temperature also influence lake level fluctuations. At higher air temperatures, plants tend to use more water, resulting in more transpiration, and there are higher rates of evaporation from both the ground surface and the lake. This yields less runoff for the same amount of precipitation than would exist during a low temperature period when there is less evaporation and transpiration. Coupled with the higher lake evaporation, lake levels drop with increasing air temperature, all other things being equal. The annual mean air temperature around the perimeter of the Great Lakes since 1900, summarized in Figure 4, indicate three distinct temperature regimes; a low temperature regime from 1900-1929 to a higher temperature regime from about 1930-1959, and an additional low regime from 1960-present period. The difference between the previous and current regime is a drop of about 1°F.

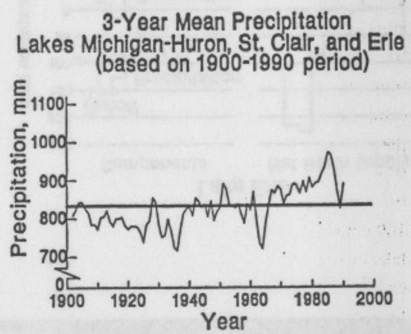


Figure 3. Historical Precipitation.

Table 3. Great Lakes Annual Precipitation Summary.

Period	Superior (cm)	Superior (in)	Michigan (cm)	Michigan (in)	Huron (cm)	Huron (in)	Erie (cm)	Erie (in)	Ontario (cm)	Ontario (in)
1900-39	72	29	78	31	77	31	85	34	86	34
1940-90	81	32	82	33	86	34	89	35	93	37
1970-90	84	33	86	34	89	35	94	37	98	39
1985[a]	105[b]	41[b]	97[b]	38[b]	106[b]	42[b]	107	42	94	37
1900-69[c]	75	30	79	31	80	32	87	34	87	34
1900-90[c]	79	31	84	33	84	33	89	35	88	35

[a]June-December 1985 provisional data from the U. S. Army Corps of Engineers.
[b]Record high for 1900-85.
[c]Long-term period averages are supplied for comparison.

The magnitude of the hydrologic variables vary with season, as shown in Figure 5 for Lake Erie (*Quinn* 1982; *Quinn and Kelley* 1983). The monthly precipitation is fairly uniformly distributed throughout the year, while the runoff has a peak during the spring which results primarily from the spring snow melt. The runoff is at a minimum in the late summer and early fall due to large evapotranspiration from the land basin. The lake evaporation reaches a minimum during the spring and gradually increases until it reaches a maximum in the late fall or early winter. The high evaporation period is due to very cold dry air passing over warm lake surfaces. The integration of these components is depicted in the net basin supply, which consists of the precipitation plus the runoff minus the evaporation. As seen from Table 2, these three components of net basin supply are all of the same order of magnitude for each lake. Annual runoff to the lake ranges from about 62 cm (24 in) for Superior to 169 cm (67 in) for Ontario, and annual lake evaporation ranges from about 56 cm (22 in) for Superior to 90 cm (35 in) for Erie. The net basin supply is seen in Figure 5 to reach a maximum in April

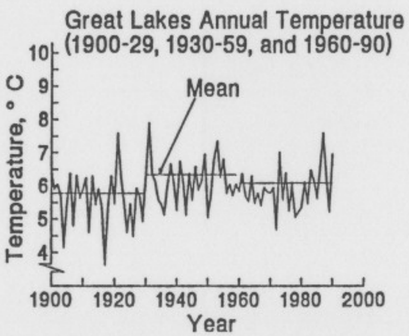


Figure 4. Historical Air Temperature.

and a minimum in the late fall. The negative values indicate that more water is leaving the lake through evaporation than is being provided by precipitation and runoff.

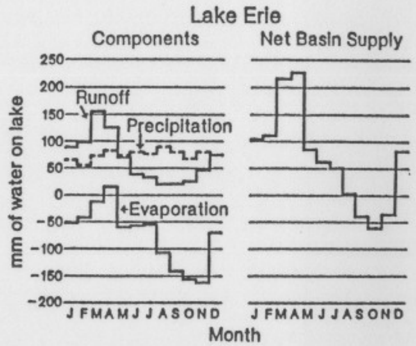


Figure 5. Seasonal Net Basin Supply.

2.4 Lake Level Fluctuation & Trends

There are three primary types of lake level fluctuations: long-term lake levels (represented on an annual basis), seasonal lake levels, and short-period lake level changes due to wind setup and storm surge. Annual fluctuations result in most of the variability leading to the record high and low lake levels. The annual lake levels are shown in Figure 6 from 1860 through the present to illustrate the long-term variability of the system. The record highs in 1952 and 1973 and record lows in 1935 and 1964 are readily apparent. There is an overall range of about 2 m (6 ft) in the annual levels. Of particular interest is the fall in the levels of Lakes Michigan and Huron occurring in the mid-1880's from which the lakes never recovered. This probably results from dredging for deeper draft navigation in the St. Clair River. Other changes in the St. Clair River include sand and gravel dredging between about 1908 and 1924, a 7.6 m (25 ft) navigational project in the mid-1930's, and an 8.2 m (27 ft) navigation project in the late 1950's and early 1960's. Without these changes, Lake Michigan-Huron would be approximately 0.5 m (1.5 ft) higher than it is today.

The three-year precipitation mean in Figure 3 correlates very well with annual lake levels as observed by superimposing the annual precipitation on the annual Lake Erie water levels in Figure 7. The precipitation tends to lead the water levels by approximately one year, as shown here by the 1929 highs, the 1935

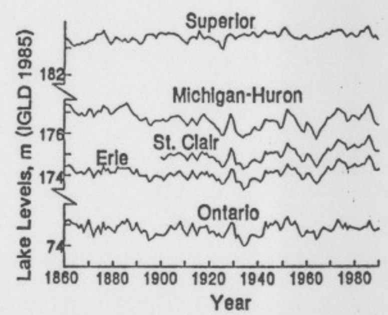
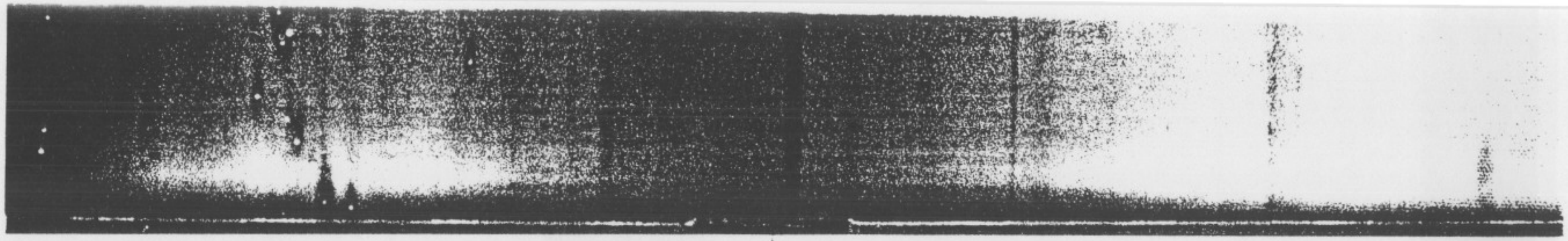


Figure 6. Historical Great Lake Levels.



lows, the 1952 highs, and the 1963 lows. In particular, the last 15 years of high precipitation resulted in very high water levels. Thus, the continuing high levels are the result of the increased precipitation regime since 1940 coupled with the lower temperature regime since 1960.

Superimposed on the annual levels are the seasonal cycles shown in Figure 8; each lake undergoes a seasonal cycle every year. The magnitude depends upon the individual water supplies. The range varies from about 30 cm (1 ft) on the upper lakes to about 38 cm (1.3 ft) on the lower lakes. In general, the seasonal cycles have a minimum in the winter, usually January or February. The levels then rise due to increasing water supplies from snow melt and spring precipitation until they reach a maximum in June for the smaller lakes, Erie and Ontario, and September in the case of Lake Superior. When the net water supplies diminish in the summer and fall, the lakes begin their seasonal decline.

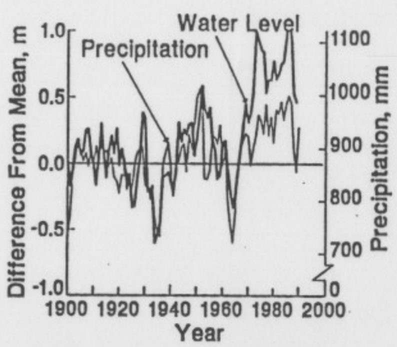


Figure 7. Annual Lake Erie Water Levels.

The final type of fluctuation which is common along the shallower areas of the Great Lakes, particularly Lake Erie, Saginaw Bay, and in some cases on Green Bay, are storm surges and wind set-up. Under these conditions when the wind is blowing along the long axis of a shallow lake or bay, a rapid difference in levels can build up between one end of the lake and the other. This difference can be as large as 5 m (16 ft) for Lake Erie (storm of 2 December 1985). These storm conditions, when superimposed on high lake levels, cause most of the damage along the Great Lakes shoreline.

Looking in more detail at the past trends in lake levels, along with the more recent conditions for Lake Erie,

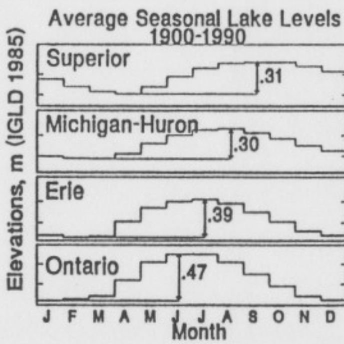


Figure 8. Seasonal Cycles.

we see a steady progression of changes in the lake levels with time in Figure 9. These changes reflect the changes in precipitation, illustrated in Figure 3 and summarized in Table 3. At the bottom of Figure 9 are the record low lake levels for each month which were set primarily in 1964. Proceeding upwards we have the 40-year average from 1900-1939. From 1940-1979, the lake is at a still higher average level. Taking the 11-year period from 1970-1980, we see that the lake level average is higher yet, followed by the record highs set in 1985. Record levels for the month were set in April and May 1985 on Lakes Michigan-Huron, St. Clair, and Erie; they were set for November 1985 through April 1986 on Lakes Erie and St. Clair. Since that time, a record drought brought water levels back to their long-term normal values in the late 1980s and early 1990s.

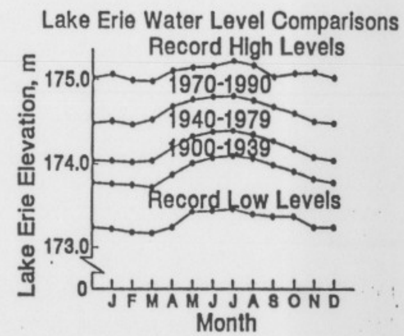


Figure 9. Lake Erie Level Comparisons.

2.5 Diversions

It is interesting to compare the impacts of the existing diversions on lake levels in Table 4 with natural lake-level fluctuations (*International Great Lakes Diversions and Consumptive Uses Study Board* 1985). This enables a comparison of man's impacts with natural fluctuations. The Long Lac and Ogoki Diversions average about 160 cms (5,600 cfs) and raise lake levels between 6 cm (0.21 ft) and 11 cm (0.37 ft). The Chicago Diversion averages about 90 cms (3,200 cfs) and lowers lake levels between 2 cm (0.07 ft) and 6 cm (0.21 ft). The Welland Canal, which bypasses Niagara Falls, averages about 270 cms (9,400 cfs) and lowers lake levels between 2 cm (0.06 ft) and 13 cm (0.44 ft) with no effect on Lake Ontario. The combined effect on the lakes ranges from a 2 cm (0.07 ft) rise for Lake Superior to a 10 cm (0.33 ft) drop for Lake Erie. The diversion effects are therefore small in comparison with the one or more meter (several foot) variation associated with short-term storm movements, the 30-38 cm (1-1.3 ft) seasonal cycle, and the 2 m (6 ft) range of annual variations.

The small effects of the diversions along with the long response time of the system illustrate why diversions are not suitable for lake regulation. Due to the large size of the Great

Table 4. Impact of Existing Diversions on Lake Levels.

Diversion	Amount		Superior		Mich-Hur		Erie		Ontario	
	(m^3s^{-1})	(cfs)	(cm)	(ft)	(cm)	(ft)	(cm)	(ft)	(cm)	(ft)
Ogoki-Long Lac	160	5600	+6	+0.21	+11	+0.37	+8	+0.25	+7	+0.22
Chicago	90	3200	-2	-0.07	-6	-0.21	-4	-0.14	-3	-0.10
Welland	270	9400	-2	-0.06	-5	-0.18	-13	-0.44	0	0
COMBINED			+2	+0.07	-1	-0.02	-10	-0.33	+2	+0.08

Lakes system, it responds very slowly to man-induced changes. This is illustrated in Figure 10 by the length of time it takes from the start of a hypothetical diversion on Lakes Michigan and Huron (of the magnitude of the Chicago diversion) until the ultimate effect of that diversion is reached on Lakes Michigan-Huron, and Erie. It takes approximately 3-3.5 years to achieve 50% of the ultimate effect and 12-15 years to get 99% of the effect. (These results depend somewhat on the lake levels at the beginning of the diversion.) Thus, regulation by diversion would not produce changes responsive to natural fluctuations. Recent studies at GLERL indicate that an increase of 10% in the Niagara River discharge from Lake Erie (and consequent increases in Lake Erie inflow) would lower it 27 cm (10.5 in) in about 11-12 years and lower Lakes Michigan and Huron 14 cm (5-6 in) in this same period. If Lake Erie inflows were held constant (not possible at the present time), then it would take 6 months to 1 year to achieve this lowering.

Additional interbasin diversions are a highly controversial issue at the present time around the Great Lakes. Possible uses of Great Lakes water outside the basin are flow augmentation for navigation, energy uses such as synthetic fuels or pipelines, agriculture and aquifer recharge, and municipal water supplies. A small pipeline project

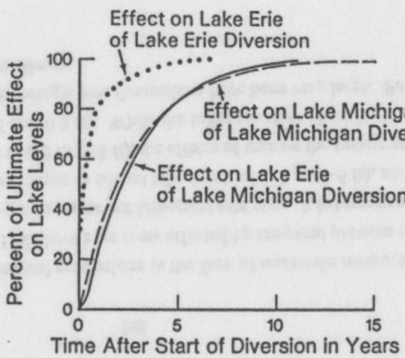


Figure 10. Response to Diversions.

such as the Powder River coal slurry pipeline would require 0.2 cms (5-8 cfs) of water and would have no measurable impact on lake levels. A synthetic fuels project, highly unlikely at this time, could require approximately 23 cms (800 cfs) and result in a lake level lowering of 1-2 cm (0.04-0.06 ft). A major agricultural or aquifer recharge project could require 300 cms (10,000 cfs) and would result in lake level decreases ranging from 12 cm (0.4 ft) on Lake Erie to 21 cm (0.7 ft) on Lake Michigan-Huron. It should be emphasized that these are hypothetical projections for illustration only.

2.6 Future

Water levels ordinarily do not change fast, as shown by the above consideration of diversions. Other studies at GLERL indicate that if normal meteorological conditions were realized ("normal" being the average conditions over 1900-69) instead of the record drought of the late 1980s, it would have taken about 6 years for Lake Michigan-Huron to return from its January 1986 level to its normal (1900-69) level. About 7 years would have been required for Lakes St. Clair and Erie to return to within 4 in of normal, and about 9 years would have been required for them to return to within 2 in of normal. Even supposing that we encountered a drought similar to the 1960-64 conditions, about 3.5 years would have been required for Lake Michigan-Huron and about 4 years would have been required for Lakes St. Clair and Erie.

A long-term perspective on Lake Michigan levels for 7,000 years was reconstructed through geologic and archaeologic evidence (*Larsen* 1985) under work sponsored by the Illinois Geologic Survey. Conditions several thousand years ago were not necessarily the same as today due to isostatic rebound and uplift during the intervening time. But, in general, this provides additional perspective on possible conditions we may experience in the future. Looking at just the last 2,500 years, during which time the Great Lakes were in their current state, there were major lake level fluctuations. During most of this time the levels were much higher and more variable than they have been during the last 120 years of record. If the past is any indication, lake levels in the future could go through a considerably larger range than we have experienced lately. Indeed, the period of record which makes up what many consider to be normal, the early 1900's through the 1960's, may be abnormal conditions.

2.7 Summary Comments on Great Lakes Dynamics

Huge storages of water in the basins and the lakes and of energy in the lakes give the Laurentian Great Lakes their characteristic behavior. They filter the variability of the meteorological inputs and enable hydrological predictions in the face of uncertain meteorology, if the storage amounts are known. Lake levels are most affected by temporal patterns of precipitation; air temperature patterns play a lesser but important role also. It is important to keep in perspective that while we have ranges in annual lake levels of 1-2 m (4-6 ft), and additional short term effects on the order of 2-3 m (7-8 ft), the effects of man on the system are relatively small, on the order of about 5 cm (0.2 ft). While the lakes are slow changing over the long term in the face of normal meteorology, past fluctuations have been very large. Future changes will depend mostly on future climate.

3. Laurentian Great Lakes Physical Process Models

GLERL developed, calibrated, and verified conceptual model-based techniques for simulating hydrological processes in the Great Lakes (including Georgian Bay and Lake St.Clair separately). GLERL integrated the models into a system to estimate lake levels, whole-lake heat storage, and water and energy balances for forecasts and for assessment of impacts associated with climate change (*Croley* 1990, 1993a,b; *Croley and Hartmann* 1987, 1989; *Croley and Lee* 1993; *Hartmann* 1990). These include models for rainfall-runoff [121 daily watershed models (*Croley* 1982, 1983a,b; *Croley and Hartmann* 1984)], over-lake precipitation (a daily estimation model), one-dimensional (depth) lake thermodynamics [7 daily models for lake surface flux, thermal structure, and heat storage (*Croley* 1989a,b, 1992a; *Croley and Assel* 1994)], channel routing [4 daily models for connecting channel flow and level, outlet works, and lake levels (*Hartmann* 1987, 1988; *Quinn* 1978)], lake regulation [a monthly plan balancing Lakes Superior, Michigan, and Huron (*International Lake Superior Board of Control* 1981, 1982) and a quarter-monthly plan balancing Lake Ontario and the St. Lawrence Seaway (*International St. Lawrence River Board of Control* 1963)], and diversions and consumption (*International Great Lakes Diversions and Consumptive Uses Study Board* 1981).

3.1 Runoff Modeling

The GLERL Large Basin Runoff Model (LBRM) consists of moisture storages arranged as a serial and parallel cascade of "tanks" (*Croley* 1983a,b); water flows from the snowpack to the upper soil zone tank, from the upper to the lower soil zone and surface storage tanks, from the lower to the groundwater and surface tanks, from the groundwater to the surface tank, and from the surface tank out of the watershed; see Figure 2. It makes use of physical concepts for snow melt and net supply to the watershed surface, infiltration, heat available for evapo-

transpiration, actual evapotranspiration, and mass conservation. As a conceptual model, the LBRM is useful not only for predicting basin runoff, but to facilitate understanding of watershed response to natural forces as well. The main mathematical feature of the LBRM is that it may be described by strictly continuous equations; none of the complexities associated with inter-tank flow rate dependence on partial filling are introduced. For a sufficiently large watershed, these nuances are not observed due to the spatial integration of rainfall, snow melt, and evapotranspiration processes.

Daily precipitation, temperature, and insolation (the latter available from climatological summaries as a function of location) may be used to determine snowpack accumulations and net surface supply based on degree-day determinations of snow melt. The net surface supply is divided into infiltration to the upper soil zone and surface runoff by taking infiltration proportional to the net surface supply rate and to the areal extent of the unsaturated portion of the upper soil zone. Outflow from each storage within the watershed is proportional to the moisture in storage. The evapotranspiration rate from the upper and lower soil zones is proportional to available moistures there and to the heat rate available for evapotranspiration; it also reduces the heat available for subsequent evapotranspiration. The total amount of heat in a day is split between that used for and that still available for evapotranspiration by empirical functions of air temperature based on a long-term heat balance. Mass continuity yields a first-order linear differential equation for each of the moisture storages (*Croley* 1982), which are tractable analytically; they are solved simultaneously to determine daily moisture storage, evapotranspiration, and basin runoff from daily data.

The Great Lakes basin is divided into 121 watersheds, each draining directly to a lake, grouped into the six lake basins. The meteorologic data from about 1,800 stations about and in the watersheds are combined through Thiessen weighting to produce areally-averaged daily time series of precipitation and maximum and minimum air temperatures for each watershed (*Croley and Hartmann* 1985b). Records for all "most-downstream" flow stations are combined by aggregating and extrapolating for ungauged areas to estimate the daily runoff to the lake from each watershed. The LBRM is calibrated to determine the set of parameters resulting in the smallest sum-of-squared-errors between model and actual daily flow volumes for the calibration period (*Croley* 1983b, *Croley and Hartmann* 1984, 1985a). After the LBRM is calibrated for each watershed, the model outflows are combined to represent each Great Lake basin; this distributed-parameter model integration filters individual sub-basin model errors. The LBRM calibration periods generally cover 1965-1982 depending upon flow data availability. Table 5 presents overall calibration results for the distributed-parameter applications. The LBRM was also used in forecasts of Lake Superior water levels (*Croley and Hartmann* 1987), and comparisons with climatic outlooks showed the runoff model was very close to actual runoff (monthly correlations of water supply were on the order of 0.99) for the period

Table 5. Large Basin Runoff Model Calibration Statistics[a].

Lake	Number of Sub-basins	Mean 1-day Flow (mm)[b]	Flow Std. Dev. (mm)[b]	Root Mean Square Error (mm)[b]	Correlation Calib.	Correlation Verif.
Superior	22	1.12	0.67	0.25	0.93	0.77
Michigan	29	0.89	0.47	0.18	0.93	0.86
Huron	27	1.06	0.69	0.26	0.92	0.69
St. Clair	7	0.90	1.36	0.62	0.89	0.87
Erie	21	1.01	1.28	0.54	0.91	0.90
Ontario	15	1.41	1.13	0.43	0.93	0.89

[a]Statistics and calibrations generally cover 1966-83; verification generally covers 1956-63.
[b]Equivalent depth over the land portion of the basin.

August 1982 - December 1984 which is outside of and wetter than the calibration period (*Croley and Hartmann* 1986). The model also was used to simulate flows for the time period 1956-63, outside of the period of calibration. The correlation of monthly flow volumes between the model and observed values during this verification period are also contained in Table 5. They are a little lower than the calibration correlations but quite good except for Lakes Superior and Huron (there were less than two-thirds as many flow gages available for 1956-63 as for the calibration period for these basins).

3.2 Over-lake Precipitation

The lack of over-lake precipitation measurements means that estimates typically depend on land-based measurements, and there may be differences between land and lake meteorology. Although gage exposures may significantly influence the results of lake-land precipitation studies (*Bolsenga* 1977, 1979), *Wilson* (1977) found that Lake Ontario precipitation estimates, based on only near-shore stations, averaged 5.6% more during the warm season and 2.1% less during the cold season than estimates based on stations situated in the lake. By using a network that also included stations somewhat removed from the Lake Ontario shoreline, *Bolsenga and Hagman* (1975) found that eliminating several gages not immediately in the vicinity of the shoreline increased over-lake precipitation estimates during the warm season and decreased them during the cold season. Thus, for the Great Lakes, where lake effects on

near-shore meteorology are significant and the drainage basins have relatively low relief, the use here of all available meteorologic stations throughout the basin is probably less biased than the use of only near-shore stations.

3.3 Over-lake Evaporation

Current Great Lakes evaporation studies use mass transfer formulations that include atmospheric stability effects on the bulk transfer coefficients, applied to monthly data for water surface temperatures, wind speed, humidity, and air temperatures (*Quinn* 1979). The present study uses that approach applied to daily data but combined with models for over-water meteorology, ice cover, and lake heat storage and with a lumped representation of a lake's heat balance (*Croley* 1989a,b, 1992a); see Figure 11. As over-water data are not generally available, over-land data are used by adjusting for over-water conditions. *Phillips and Irbe's* (1978) regressions for over-water corrections are used directly by replacing the fetch (and derived quantities) with averages. Air temperatures and specific humidities over ice are used for over-ice evaporation calculations and over water for the over-water calculations; the two estimates are combined by weighting for the fraction of the surface covered in ice. Water and ice pack heat balances (*Croley and Assel* 1994) were used to relate ice cover extent to meteorology, heat storage, and surface fluxes between the atmosphere, the water body, and the ice pack.

Kraus and Turner's (1967) mixed-layer thermal structure concept is extended for the Great Lakes to allow the determination of a simple one-dimensional model for surface temperature increments or decrements from past heat additions or losses, respectively (*Croley* 1989a,b, 1992a). The effects of past additions or losses are superimposed to determine the surface temperature on any day as a function of heat in storage; each past addition or loss is parameterized by its age. Turnovers (convective mixing of deep lower-density waters with surface waters as surface temperature passes through that at maximum density) can occur as a fundamental behavior of this superposition model, and hysteresis between heat in storage and surface temperature, observed during the heating and cooling cycles on the lakes, is preserved.

Heat in storage in the lake at the end of each day is given by a simple conservation of energy by taking the change in storage equal to the sum of the fluxes integrated over the day. As summarized by *Gray et al.* (1973), short-wave radiation is interpolated from generalized maps of Canadian and northern U.S. mid-monthly clear-sky values and adjusted for cloud cover. Average short-wave reflection is taken simply as one-tenth of the incident or as a function of ice cover, and sensible heat transfers at the water or ice temperature (minimum of air temperature or freezing temperature) are computed directly from the same mass transfer formulation and assumption (that the bulk evaporation coefficient is equal to the sensible heat coefficient) that is used to derive evaporation. It is then added to evaporative advection and latent heat transfers. Evaporative heat transfers from ice include the heat of fusion as well. Net long-wave radiation exchange is derived from considerations of the water and atmosphere as gray bodies with correction for cloud cover only to atmospheric radiation (*Keijman* 1974). Net long-wave radiation exchange over ice is computed as for open water, ignoring the small effects of the ice surface on the exchange. Energy advected with precipitation is adjusted if the precipitation is snow, to account for the heat required in snow melt. Energy advected with precipitation onto the ice surface is uncorrected for melt since that is taken as occurring with ice melt, which is added to the budget when it happens. The energies advected into and out of the lake with other mass flows are relatively very small and are ignored. The equations representing evaporation, heat storage, and heat fluxes are solved simultaneously with daily data on over-land wind speed, air temperature, cloud cover, and humidity; details of an iterative solution technique are available elsewhere (*Croley* 1989a,b, 1992a).

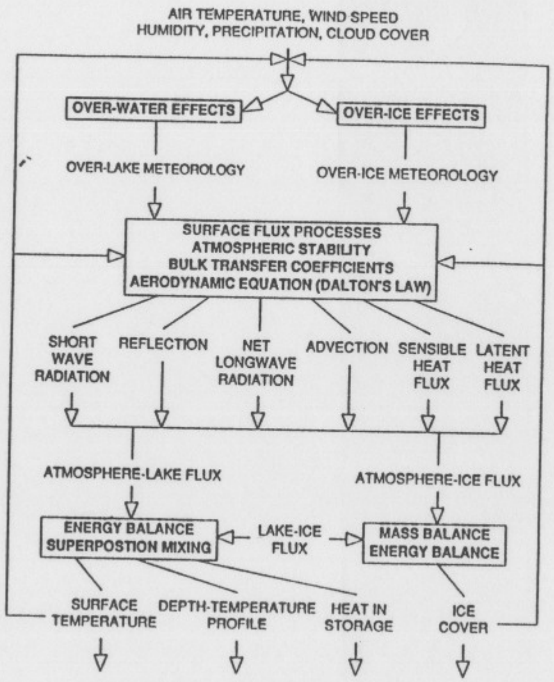


Figure 11. Evaporation Model Schematic

Unfortunately, there are no really good independent evaporation data to calibrate and verify evaporation models on the Great Lakes. Water balances are insufficient due to the large errors induced by subtracting nearly equal large inflows and outflows to each Great Lake, or due to errors in estimates of the water balance components. However, with the joint heat balance and evaporation model described here, it is possible to compare water surface temperatures with data, now available from the National Oceanic and Atmospheric Administration's Polar Orbiting Satellite Advanced Very High Resolution Radiometer (*Irbe et al.* 1982; *AES* 1988).

Daily meteorological over-land data at from five to seven near-shore stations about each Great Lake were assembled and averaged for correction to over-lake data. The heat balance model was calibrated to give the smallest sum-of-squared-errors between model and actual daily water surface temperatures observed by satellite during the calibration period of generally 1979-88; the results are summarized in Table 6. There is good agreement between the actual and calibrated-model water surface temperatures; the root mean square error is between 1.1-1.6°C on the large lakes [within 1.1-1.9°C for an independent verification period, 1966-78 (*Croley* 1989a,b, 1992a)].

3.4 Models Validity

Although GLERL uses a daily resolution of data with their models, basin-wide processes of runoff, over-lake precipitation, and lake evaporation (described with models here) respond discernibly to weekly changes at best, and monthly is usually adequate for net supply and lake level simulation (this ignores short-term fluctuations associated with storm movement which are not addressed in this study). Likewise, spatial resolution finer than about 1,000-5,000 km^2 (the present average resolution of GLERL's models and their applications) is unnecessary and much can be done in assessing hydrology changes at resolutions of 100,000 - 1,000,000 km^2 with lumped versions of the models. This coarse spatial resolution is still much finer than present general circulation model (GCM) grids.

The models were partially assessed by computing net basin supplies to the lakes with historical meteorological data for 1951-80 and comparing to historical net basin supplies. The absolute average annual difference ranged from 1.6% to 2.7% on the deep lakes while the Lake St. Clair and Lake Erie applications were 12.0% and 7.0% respectively; month-to-month differences showed more variation. These differences generally reflect poorer evaporation modeling on the shallow lakes and snow melt and evapotranspiration model discrepancies for the other lake basins. While monthly differences were generally small, a few were significant. The low annual residuals were felt to be acceptable to use these models in assessing changes from the current climate as they would be consistently applied to both a "present" and all budget term errors in the derived net basin supplies.

Table 6. Daily Lake Evaporation Model Calibration Results.

	Superior	Michigan	Huron	Georgian	Erie	Ontario
CALIBRATION PERIOD STATISTICS						
Water Surface Temperatures (1980-88)[a]						
Means Ratio[b]	1.00	1.01	0.98	1.01	1.03	0.99
Variances Ratio[c]	1.01	0.98	0.95	1.02	1.08	0.99
Correlation[d]	0.98	0.97	0.98	0.99	0.99	0.98
R. M. S. E.[e]	1.13	1.56	1.33	1.10	1.58	1.43
Ice Concentrations (1960-1988)[f]						
Means Ratio[g]	0.92	0.72	0.70	0.98	1.15	0.39
Variances Ratio[h]	1.24	1.02	1.67	1.62	1.09	0.63
Correlation[i]	0.76	0.83	0.73	0.77	0.89	0.54
R. M. S. E.[j]	23.4	12.4	26.0	21.5	19.0	15.4
VERIFICATION PERIOD STATISTICS						
Water Surface Temperatures (1966-79)[k]						
Means Ratio[b]	0.96		1.03	0.98	1.05	0.94
Variances Ratio[c]	1.10		0.95	1.00	1.10	0.97
Correlation[d]	0.97		0.99	0.98	0.98	0.96
R. M. S. E.[e]	1.09		1.10	1.34	1.91	1.92

[a]Data between January 1, 1980 and August 31, 1988 for all lakes except Michigan and between January 1, 1981 and August 31, 1988 for Lake Michigan, with an initialization period for all lakes except Georgian Bay starting January 1, 1948 and January 1, 1953 for Georgian Bay.
[b]Ratio of mean model surface temperature to data mean.
[c]Ratio of variance of model surface temperature to data variance.
[d]Correlation between model and data surface temperature.
[e]Root-mean-square error between model and data surface temperatures in degrees C.
[f]Data between January 1, 1960 and August 31, 1988 for all Great Lakes except Superior and between March 1, 1963 and August 31, 1988 for Lake Superior, with an initialization period for all lakes starting January 1, 1958.
[g]Ratio of mean model ice concentration to data mean.
[h]Ratio of variance of model ice concentration to data variance.
[i]Correlation between model and data ice concentration.
[j]Root-mean-square error between model and data ice concentrations in %.
[k]Data between January 1, 1966 and December 31, 1979 for all lakes except Michigan with an initialization period for all lakes except Georgian Bay starting January 1, 1948 and January 1, 1953 for Georgian Bay.

a "changed" climate. Further assessment of model deficiencies with comparisons to historical net basin supplies is difficult since the latter are derived from water budgets which incorporate

There is some indication of model applicability outside of the time periods over which the models were calibrated as indicated above and in Tables 5 and 6. To assess the applicability of the process models to a climate warmer than the one under which they were calibrated and verified requires access to meteorologic data and process outputs for the warmer climate which unfortunately do not exist. Warm periods early in this century are not sufficiently documented for the Great Lakes. In particular, data are lacking on watershed runoff to the lakes, water surface temperatures, wind speed, humidity, cloud cover, and solar insolation.

It is entirely possible that the models are tied somewhat to the present climate; empiricism is employed in the evapotranspiration component of the LBRM and in some of the heat flux terms in the heat balance and lake evaporation model. Coefficients were determined or selected in accordance with the present climate. The models are all based on physical concepts that should be good under any climate; however, the assumption is made that they represent processes under a changed climate that are the same as the present ones. These include linear reservoir moisture storages, partial-area infiltration, lake heat-storage relations with surface temperature, and gray-body radiation. However, the calibration and verification periods for the component process models include a range of air temperatures, precipitation, and other meteorological variables that encompass much of the changes in these variables predicted for a changed climate. Even though the changes are transitory in the calibration and verification period data sets, the models appear to work well under these conditions.

4. Laurentian Great Lakes Climate Change Response

Climatic change will impact many aspects of the hydrologic cycle with interrelated consequences for mankind. A doubling of atmospheric CO_2 will impact Great Lakes water supply components and basin storages of water and heat that must be understood before lake level impacts can be assessed. Because the Laurentian Great Lakes possess tremendous water and heat storage capacities, they respond slowly to changed meteorologic inputs. This "memory" results in a filtering or dampening of most short-term meteorologic fluctuations and in a response to longer-period fluctuations characteristic of climate change. The large Great Lakes system thus is ideal for studying regional effects of climate changes.

Preliminary estimates of the impact of climatic warming on Great Lakes water resources are summarized elsewhere (*Croley and Hartmann* 1989; *Croley* 1990). The Environmental Protection Agency (EPA) coordinated several regional studies of various impacts of a doubling of atmospheric CO_2 at the direction of the U.S. Congress. As part of that study, GLERL assessed steady-state and transient changes in Great Lakes hydrology consequent with simulated atmospheric scenarios from three GCMs. Those studies, in part, and the high water levels of the late 1980s prompted the International Joint Commission to reassess climate change impacts on Great Lakes hydrology and lake thermal structure.

The methodology established in the EPA studies was adopted with slight modifications for use in the IJC studies. The methodology integrates hydrology and lake heat storage models (*Croley* 1991, 1992b) to consider climate scenarios supplied by the Canadian Climate Centre (CCC) from its GCM (*Louie* 1991; *McFarlane* 1991). *Cohen* (1991) discusses the problems with this approach. The CCC provided a "present-climate" meteorology simulation ($1xCO_2$) and a changed-climate scenario ($2xCO_2$) developed from their atmospheric global circulation model. GLERL abstracted differences between the CCC-generated $1xCO_2$ and $2xCO_2$ atmospheres, made these changes to historical data, and observed the impact of the changed data in the hydrological outputs of their models.

The EPA studies included partial assessments of large-lake heat storage associated with climate change on Lakes Michigan (*McCormick* 1989) and Erie (*Blumberg and DiToro* 1989). The IJC study looked in less detail but more breadth at large-lake thermodynamics in that while only lake-wide effects were considered, all lakes were assessed. This section presents the methodology of linkage between regional hydrological models and the GCMs, describes their limitations, and presents and interprets the IJC studies of hydrological changes predicted through use of the Canadian Climate Centre's GCM.

4.1 Methodology

GLERL constructed a master computer procedure to integrate the Large Basin Runoff Model, over-lake precipitation estimates, and the lake evaporation models for all lakes to provide a net water supply model for the entire Great Lakes system. They developed it specifically to look at the impact of changed climate by doing simulations with changed meteorology that represent scenarios of changed climate and comparing with simulations based on historical meteorology (representing an unchanged climate). Inputs are areal-average daily precipitation and maximum and minimum air temperatures for each of the 121 watersheds about the Great Lakes and areal-average daily air temperature, cloud cover, humidity, and wind speed for each of the five Great Lakes and Lake St. Clair.

GLERL's general procedure for the investigation of steady-state behavior under a changed climate is similar to that used for the EPA, as detailed elsewhere (*Croley* 1991; *Louie* 1991); it required that GLERL first simulate 38 years of "present" hydrology by using historical daily maximum and minimum air temperatures, precipitation, wind speed, humidity, and cloud cover data for the 1951-88 period; this is called the "base case" or "$1xCO_2$" scenario. The initial conditions were arbitrarily set, but an initialization simulation period of 1 January 1948 through 31 December 1950 was used to allow the models to converge to conditions (basin moisture storages, water surface temperatures, and lake heat storages) initial to the 1 January 1951 through 31 December 1988 period. GLERL then attempted to estimate "steady-state" conditions, but there were problems.

The procedure to estimate "steady-state" conditions is to repeat the 41-year simulation with initial conditions (basin moisture storages, lake heat storages, and surface temperatures) set equal to their values at the end of the simulation period, until they are unchanging. This procedure requires many iterations for a few sub-basins with very slow groundwater storages and suggests very different initial groundwater storages than were used in calibrations. Actually, the original calibrations of the models used arbitrary (but fixed) initial conditions. GLERL should have determined initial conditions in the calibrations, but that was unfeasible; there is little confidence in calibrated parameter sets that suggest very slow groundwater storages (half-lives on the order of several hundred years in some cases) since only 10-20 years were used in the calibrations. Therefore, the best estimate of "present" hydrology is to use calibrated parameters with initial conditions on "the same order" as those assumed for the calibrations. GLERL did the latter and then conducted simulations with adjusted data sets.

Average monthly absolute air temperatures, specific humidities, cloud cover, precipitation, and wind speed were supplied for each month of the year by the Canadian Climate Centre as resulting from their second-generation global circulation model; see *McFarlane* (1991). While available at grid points spaced 3.75 degrees latitude by 3.75 degrees longitude, *Louie* (1991) interpolated monthly averaged data to a grid of 1 degree latitude by 1 degree longitude for both the "present" and "future" atmospheres (with one and two times the CO_2 content of the "present" atmosphere). He weighted values at surrounding grid points inversely to the square of the distance to each point. GLERL computed ratios of "future" ($2xCO_2$) to "present" ($1xCO_2$) monthly average absolute air temperatures, specific humidities, cloud cover, and precipitation and monthly average differences of $2xCO_2$ to $1xCO_2$ wind speeds at each of these grid points. They then used these ratios and differences with the historical data to estimate the 41-year sequences (1948-88) of atmospheric conditions associated with a changed climate, referred to as the "$2xCO_2$" scenario.

GLERL inspected each of the 770,000 square kilometers within the Great Lakes Basin to see which grid point it is closest to and applied the monthly adjustment at that grid point to data representing that square kilometer. By combining all square kilometers representing a watershed or the lake surface, GLERL derived an areally-averaged adjustment to apply to their areally-averaged data sets for the watershed or lake surface, respectively. They then used the $2xCO_2$ scenario in simulations similar to the base case scenario. They repeated the

41-year simulation with initial conditions set equal to their values at the end of the simulation period, until they were unchanging to estimate "steady-state" future conditions. They then interpreted differences between the 2xCO_2 scenario and the base case scenario, for the 1951-88 period, as resulting from the changed climate.

Transfer of information between the GCMs and GLERL's hydrologic models in the manner described involves several assumptions. Solar insolation at the top of and through the atmosphere on a clear day are assumed to be unchanged under the changed climate, modified only by cloud cover changes. Over-water corrections are made in the same way, albeit with changed meteorology, which presumes that over-water/over-land atmospheric relationships are unchanged. GLERL's procedure for transferring information from the GCM grid to their spatial data is an objective approach but simple in concept. It ignores interdependencies in the various meteorologic variables as all are averaged in the same manner. Of secondary importance, the spatial averaging of meteorologic values over a box centered on the GCM grid point (implicit in the use of the nearest grid point to each square kilometer of interest) filters all variability that exists in the GCM output over that box. If GCM output was interpolated between these point values, then at least some of the spatial variability might be preserved. The interpolation performed by *Louie* (1991) from the original GCM grid to a finer grid reduced this problem, but it still exists in the use of the finer grid with the hydrology models. Of course, little is known about the validity of various spatial interpolation schemes and, for highly variable spatial data, they may be inappropriate. However, the same is true for the spatial averaging that was used to supply the GCM results for this study.

Steady-state behavior, in all aspects of the hydrological cycle, are exemplified here in figures for the Lake Superior basin and summarized for all lakes and all climate-change scenarios for the entire period in tables.

4.2 Basin Meteorology

The annual cycles, of all meteorologic variables, were averaged over the 1951-88 period and inspected. The 2xCO_2 climate air temperatures are higher throughout the annual cycle than the 1xCO_2 climate (base case); the difference is smallest during the late fall to early winter and largest during the late winter to early spring for all lakes; see Figure 12. The difference is smallest and largest (most variability in the seasonal cycle) for the southern-most lakes. The average annual air temperatures are 4.4-6.1°C higher, depending on the basin; see Table 7. The 2xCO_2 climate precipitation is generally higher during the spring and lower during the fall and winter than the 1xCO_2 climate precipitation over all of the Great Lake basins, although generally lower to the south; see Figure 12. The average steady-state annual precipitation is 8% higher over the Superior basin to 10% lower over the Erie basin with a fairly smooth change with longitude; see Table 7. Precipitation changes and air temperature changes are both fairly consistent with longitude as illustrated in Table 7.

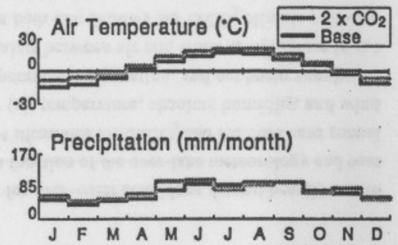


Figure 12. Over-Basin Meteorology Changes.

4.3 Basin Hydrology

The resulting average annual steady-state evapotranspiration from the land portion of the basins is higher for the 2xCO_2 climate in all lake basins, with a fairly smooth change with longitude from 26% higher over the Superior

Table 7. Average Annual Steady-State Basin Hydrology Differences.

Basin	Air Temperatures and Absolute Differences: 1xCO_2	2xCO_2	Diff.	Precipitation[a] and Relative Changes: 1xCO_2	2xCO_2	Diff.
Superior	2.4°	6.8°	4.4°	817 mm	880 mm	8 %
Michigan	7.2°	12.8°	5.6°	825 mm	797 mm	-3 %
Huron	5.4°	10.4°	5.0°	870 mm	852 mm	-2 %
St. Clair	8.3°	14.4°	6.1°	849 mm	772 mm	-9 %
Erie	9.1°	15.1°	6.0°	905 mm	817 mm	-10 %
Ontario	7.2°	12.2°	5.0°	930 mm	879 mm	-6 %

Basin	Evapotranspiration[a] and Relative Changes: 1xCO_2	2xCO_2	Diff.	Runoff[a] and Relative Changes: 1xCO_2	2xCO_2	Diff.
Superior	423 mm	534 mm	26 %	394 mm	346 mm	-12 %
Michigan	507 mm	600 mm	18 %	317 mm	196 mm	-38 %
Huron	493 mm	608 mm	24 %	377 mm	243 mm	-36 %
St. Clair	535 mm	632 mm	18 %	315 mm	140 mm	-56 %
Erie	565 mm	659 mm	17 %	341 mm	158 mm	-54 %
Ontario	472 mm	575 mm	22 %	459 mm	304 mm	-34 %

[a]Expressed as a depth over the land portion of the basin.

basin to 17% higher over the Erie basin; see Table 7. However, over the seasonal cycle, $2xCO_2$ evapotranspiration exceeds the base case most often in the late spring to early summer (late on Lake Superior basin) and is actually smaller in the early fall, see Figure 13. Runoff from the land portion of the basin is reduced by the $2xCO_2$ climate in all basins, changing from only 12% lower over the Superior basin to 56% lower over the St. Clair basin in a fairly smooth variation with latitude; see Table 7. The average annual cycle of runoff, depicted in Figure 13, has changed as well; runoff peaks slightly earlier and with smaller magnitude under the $2xCO_2$ climate than under the $1xCO_2$ climate. This results largely from big changes in snowpack accumulation and ablation, and in other moisture storages.

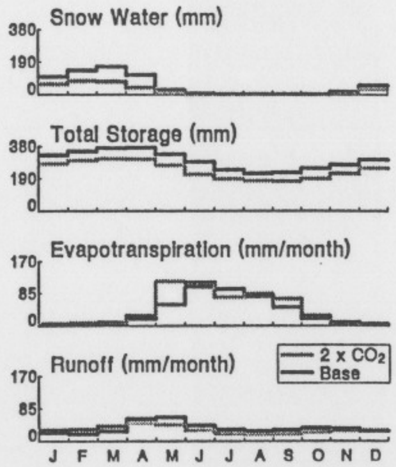


Figure 13. Over-Basin Hydrology Changes.

On the Superior basin, the average steady-state snowpack storage is reduced by more than half; on the other basins, more to the south, the snowpack is almost entirely absent under the $2xCO_2$ climate; see Figure 13 and Table 8. This reduction in snowpack accumulation results from the higher air temperatures, especially during the winter, that accompany the changed climate. The snow season is shortened more than one month. The effects on the snowpack are felt throughout the basin in terms of the derived moisture storages in the soil zone, groundwater, and surface zones. Figure 13 illustrates the general impact on all Great Lake basins of generally lower moisture storages that peak earlier in the $2xCO_2$ climate than in the $1xCO_2$ climate scenarios. This general lowering of moisture in storage in each of the basins is summarized in Table 8 and in some cases represents greater than a 50% reduction in available moisture (see "Total Basin Storage" column).

4.4 Over-Water Meteorology

The over-lake air temperature, humidity, and wind speed differs from over-land since the lower atmospheric layer is affected by the water surface over which it lies. The model correc-

Table 8. Average Annual Steady-State Basin Storage Differences.

Basin	Snow Water Equivalent[a] and Relative Changes			Soil Moisture[a] and Relative Changes		
	$1xCO_2$	$2xCO_2$	Diff.	$1xCO_2$	$2xCO_2$	Diff.
Superior	50 mm	24 mm	-51 %	42 mm	36 mm	-14 %
Michigan	12 mm	2 mm	-87 %	35 mm	22 mm	-37 %
Huron	28 mm	6 mm	-79 %	54 mm	40 mm	-26 %
St. Clair	9 mm	1 mm	-91 %	6 mm	2 mm	-67 %
Erie	6 mm	1 mm	-90 %	7 mm	3 mm	-63 %
Ontario	16 mm	2 mm	-85 %	21 mm	14 mm	-31 %

Basin	Groundwater Moisture[a] and Relative Changes			Total Basin Storage[a] and Relative Changes		
	$1xCO_2$	$2xCO_2$	Diff.	$1xCO_2$	$2xCO_2$	Diff.
Superior	146 mm	124 mm	-15 %	295 mm	237 mm	-20 %
Michigan	61 mm	41 mm	-33 %	114 mm	68 mm	-40 %
Huron	8 mm	5 mm	-39 %	99 mm	57 mm	-43 %
St. Clair	10 mm	5 mm	-51 %	28 mm	9 mm	-67 %
Erie	9 mm	4 mm	-52 %	24 mm	8 mm	-65 %
Ontario	11 mm	7 mm	-36 %	61 mm	33 mm	-46 %

[a]Expressed as a depth over the land portion of the basin.

tions to over-land meteorology observations for over-water conditions depend heavily on the water surface temperature which in turn is a function of the over-lake meteorology and heat balance at the surface of the lake. Figure 14 illustrates the $2xCO_2$ and the base case annual cycles for Superior over-lake meteorology (air temperature, absolute humidity, and wind speed) and Figure 15 illustrates water temperature, evaporation, and net basin supply. In general, the synergistic relationship that exists between air and water temperature in the $2xCO_2$ scenario yields a general increase in both that follows the $1xCO_2$ climate patterns, similar to over-land behavior in Figures 14 and 15. The most pronounced increase in both occurs in the summer for Lake Superior. Table 9 shows that the average steady-state air temperature difference between the $2xCO_2$ and base cases varies from 5.3°C on Lakes Superior and Huron to 5.9°C on Lake Michigan. Variations in the impact with latitude or longitude or size of the lake are not pronounced, in terms of volume or heat capacity. Relative humidity over the lakes is increased, probably due to the increased lake evaporation, and cloud cover generally has decreased slightly for the $2xCO_2$ climate; see Figure 14 and Table 9. Again, the difference is most pronounced in the summer. Over-water wind speed is not greatly affected after correction of over-land values for over-water conditions at increased water temperatures; Figure 14 and Table 9 show only slight decreases for each lake in average steady-state wind speed.

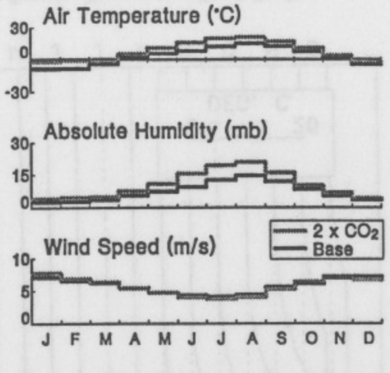


Figure 14. Over-Lake Meteorology Changes.

4.5 Lake Heat Balance

The heat budget gives rise to increased water surface temperatures as seen in Figure 15 and summarized in Table 10. The average steady-state increase in water surface temperatures for the $2xCO_2$ scenario range from 4.8°C on Lake St. Clair to 5.6°C on Lake Michigan. The heat storage capacity of a lake influences the increase in water surface temperatures that can almost be seen in Figure 15. Water surface temperatures are seen to peak earlier on deep lakes under the $2xCO_2$ climate than under the $1xCO_2$ climate. Large amounts of heat now reside in the deep lakes throughout the year, increasing latent and sensible transfers to the atmosphere; see Figure 15. The increased heat in storage also means that ice formation will be greatly reduced over winter on the deep Great Lakes. The higher water surface temperatures under the $2xCO_2$ climate result in increased annual lake evaporation of about 31-33%; see Table 10. The increased heat storage also changes the temperature-depth profiles within the lakes.

Some of the deep lakes (Michigan, Huron, and Ontario) show water surface temperatures that stay above 3.98°C throughout the average annual

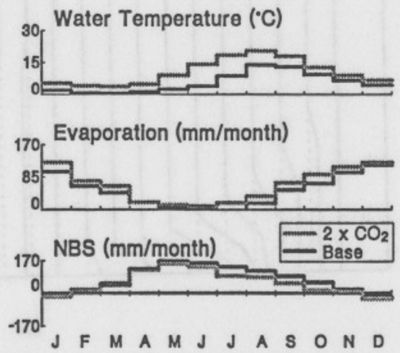


Figure 15. Over-Lake Hydrology Changes.

Table 9. Average Annual Steady-State Over-Lake Meteorology Differences.

Basin	Air Temperatures and Absolute Differences			Absolute Humidity and Absolute Differences		
	$1xCO_2$	$2xCO_2$	Diff.	$1xCO_2$	$2xCO_2$	Diff.
Superior	3.1°	8.4°	5.3°	7.2 mb	10.3 mb	3.1 mb
Michigan	7.7°	13.6°	5.9°	9.7 mb	13.3 mb	3.6 mb
Huron	6.5°	11.8°	5.3°	9.0 mb	12.4 mb	3.4 mb
St. Clair	10.3°	16.0°	5.7°	11.2 mb	15.1 mb	3.9 mb
Erie	9.7°	15.3°	5.6°	10.9 mb	14.7 mb	3.8 mb
Ontario	8.0°	13.4°	5.4°	9.8 mb	13.6 mb	3.8 mb
Basin	**Cloud Cover and Relative Changes**			**Wind Speed and Relative Changes**		
	$1xCO_2$	$2xCO_2$	Diff.	$1xCO_2$	$2xCO_2$	Diff.
Superior	0.57	0.58	2 %	5.7 m/s	5.7 m/s	0 %
Michigan	0.43	0.42	-2 %	6.1 m/s	5.9 m/s	-3 %
Huron	0.55	0.55	-1 %	6.0 m/s	5.9 m/s	-2 %
St. Clair	0.50	0.47	-4 %	5.7 m/s	5.6 m/s	-2 %
Erie	0.63	0.61	-3 %	6.3 m/s	6.1 m/s	-3 %
Ontario	0.59	0.58	-1 %	6.1 m/s	6.0 m/s	-2 %

Table 10. Selected Average Annual Steady-State Hydrology Differences.

Basin	Water Temperature and Absolute Differences			Over-Lake Evaporation[a] and Relative Changes		
	$1xCO_2$	$2xCO_2$	Diff.	$1xCO_2$	$2xCO_2$	Diff.
Superior	5.4°	10.5°	5.1°	561 mm	736 mm	31 %
Michigan	8.5°	14.1°	5.6°	647 mm	854 mm	32 %
Huron	8.0°	13.0°	5.0°	627 mm	829 mm	32 %
St. Clair	1.0°	15.8°	4.8°	936 mm	1234 mm	32 %
Erie	10.9°	15.8°	4.9°	898 mm	1197 mm	33 %
Ontario	9.0°	14.4°	5.4°	665 mm	874 mm	31 %

[a]Expressed as depths over the lake.

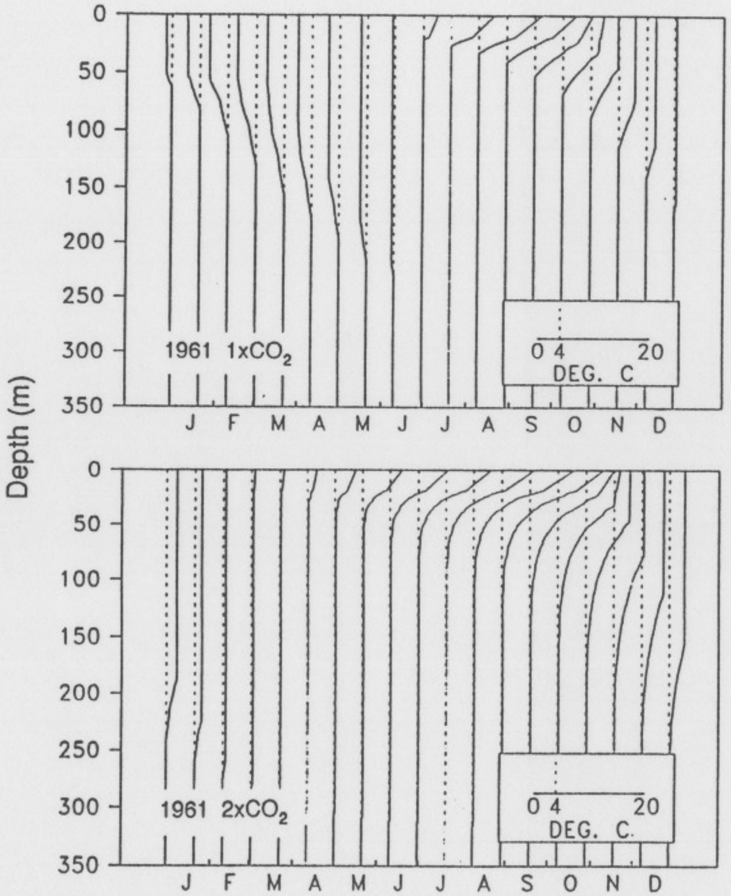


Figure 16. Steady-State Lake Superior Temperature-Depth Profiles.

cycle. Figure 16 illustrates this for 1961 on Lake Superior. This means that buoyancy-driven turnovers of the water column do not occur in the same way as they do at present.

In some years, the large lakes are changed from dimictic lakes (turnovers occur twice a year as water temperatures pass through the point of maximum density, 3.98°C) to monomictic lakes (maximum turnover occurs at the temperature "reversal" where temperatures stop declining and start rising again and the minimum temperature is greater than 3.98°). Figure 16 shows that the $1xCO_2$ temperature profile for Lake Superior passed through 3.98°C in June 1961 and approached, in December, the January 1962 transition. Under the $2xCO_2$ scenario, temperatures remain above 3.98°C but approach a vertical profile most in March. This represents a change from dimictic to monomictic.

Table 11 shows that the large lakes remain dimictic under the $2xCO_2$ climate only between 2% and 76% of the time. The largest change is associated with Lake Ontario which is the furthest south of the deep lakes. Least effected are Lakes Erie and St. Clair which are very shallow and have relatively little heat storage. As the lakes move to one reversal per year in some years, instead of two turnovers per year, the interarrival times of the maximum mixing extent increase. Table 11 illustrates that the average interarrival time grows to nearly a full year on Lake Ontario since only 2% of the years have dimictic behavior. Table 11 also illustrates the monomictic reversal temperature is, of course, well above the point of maximum water density.

The timing of maximum turnovers or temperature reversals shifts. Table 12 shows the time between the spring turnover and the fall turnover (for dimictic behavior) increases. The spring turnover occurs earlier and the fall turnover occurs later in the annual cycle. For monomictic behavior, the single maximum turnover occurs even earlier in the year than the dimictic turnovers. These are consequences of greater heat storage in, and heat inputs to, the lakes.

Temperature-depth profiles for every day of a single model year can be combined and depicted as depth-time plots of temperature isolines; see Figure 17 for an example on Lake Superior. Then, not only are the turnover timing changes depicted between $1xCO_2$ and $2xCO_2$ climates, but depth changes are more apparent as well. Table 12 also summarizes the maximum depths at turnover in the lakes. Dimictic spring turnovers exhibit shallower average depths under $2xCO_2$ conditions and fall turnovers are deeper, where not limited by the depth of the lake. Monomictic turnovers are generally even deeper.

There is a normal hysteresis observed in graphs of lake heat plotted with surface temperature, such as in Figure 18. This reflects the mixing of heat at depth. Surface temperatures rise quickly and heat storage follows after the spring turnover. When surface temperatures then begin to drop in the fall, stored heat does not initially. Then heat storage drops more slowly. Similar behavior occurs after the fall turnover, and both result in the characteristic double "loop" in the plot. Under the warmer climate change scenario, temperatures sometimes never drop below that at maximum density (3.98°C). This results in only one hysteresis loop, but it is much larger.

Table 11. Average Characteristics of Turnovers/Reversals

	Fraction Dimictic		Interarrival Times		$2xCO_2$ Monomictic Reversal
	$1xCO_2$	$2xCO_2$	$1xCO_2$	$2xCO_2$	Water Temperature
Superior	100 %	67 %	182 d	211 d	4.5 °C
Michigan	100 %	15 %	182 d	318 d	4.9 °C
Huron	100 %	24 %	182 d	292 d	4.9 °C
St. Clair	100 %	100 %	183 d	185 d	
Erie	100 %	76 %	183 d	206 d	4.9 °C
Ontario	100 %	2 %	182 d	356 d	5.9 °C

Table 12. Average Dates and Depths of Maximum Turnover or Temperature Reversal.

Basin	$1xCO_2$		$2xCO_2$		
	Dimictic		Dimictic		Monomictic
	Spring	Fall	Spring	Fall	
DATES					
Superior	02 Jul	23 Dec	24 Apr	11 Feb	24 Mar
Michigan	27 May	01 Jan	27 Mar	28 Jan	22 Feb
Huron	26 May	10 Jan	26 Mar	11 Feb	10 Mar
St. Clair	30 Apr	20 Nov	04 Mar	28 Nov	
Erie	30 Apr	24 Dec	03 Mar	06 Jan	01 Feb
Ontario	20 May	18 Jan	10 Mar	29 Jan	03 Mar
DEPTHS					
Superior	234 m	162 m	127 m	257 m	297 m
Michigan	132 m	111 m	52 m	199 m	231 m
Huron	229 m[a]	229 m[a]	140 m	229 m[a]	229 m[a]
St. Clair	6 m[a]	6 m[a]	6 m[a]	6 m[a]	
Erie	64 m[a]	64 m[a]	48 m	64 m[a]	53 m
Ontario	232 m	242 m	70 m	244 m[a]	244 m[a]

[a]Maximum average depth of the lake.

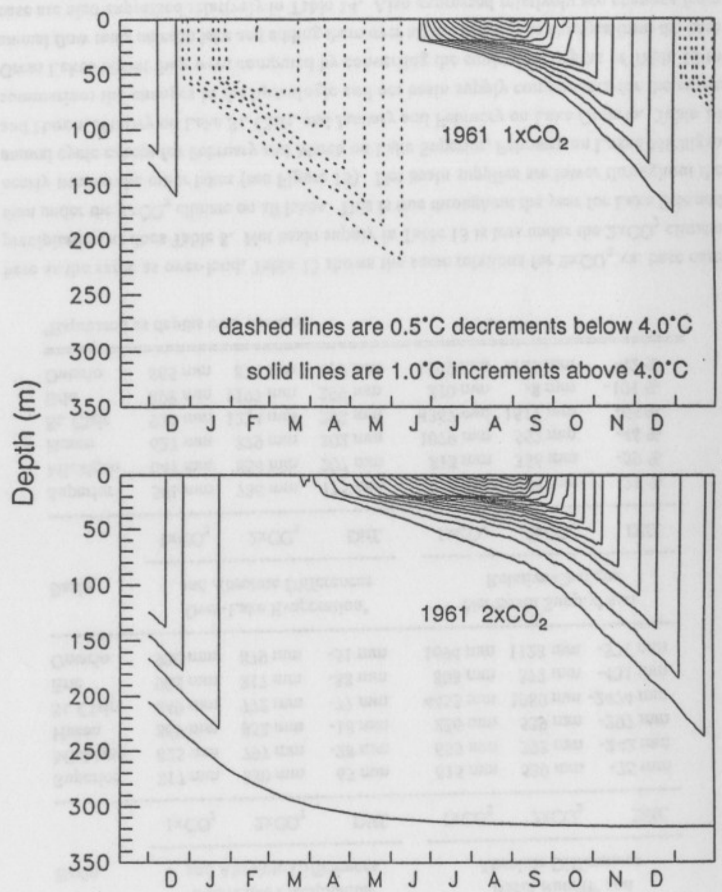


Figure 17. Steady-State Lake Superior Depth-Time Temperature Isolines.

4.6 Net Supply Components

Over-lake precipitation, runoff, and lake evaporation sum algebraically as the net basin supply and are presented again in Table 13 for convenience. Since over-lake precipitation is taken

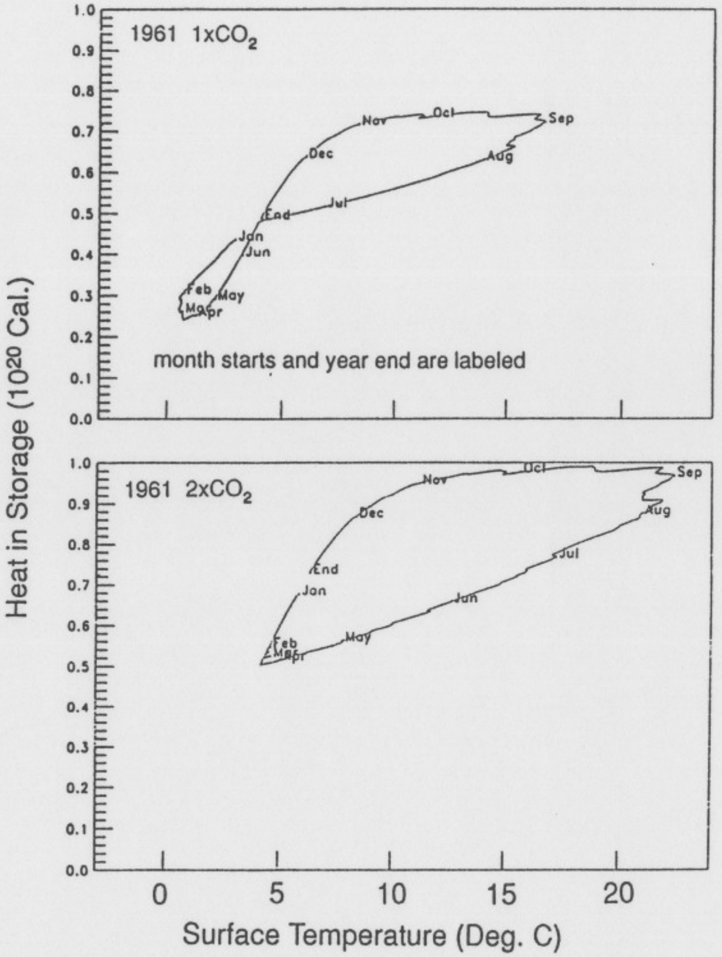


Fig. 18. Steady-State Lake Superior Heat-Temperature Hysteresis.

Table 13. Average Annual Steady-State Net Supply Components Differences.

Basin	Over-Lake Precipitation[a] and Absolute Differences			Basin Runoff[a] and Absolute Differences		
	$1xCO_2$	$2xCO_2$	Diff.	$1xCO_2$	$2xCO_2$	Diff.
Superior	817 mm	880 mm	63 mm	615 mm	539 mm	-75 mm
Michigan	825 mm	797 mm	-28 mm	635 mm	393 mm	-242 mm
Huron	869 mm	852 mm	-18 mm	836 mm	539 mm	-297 mm
St. Clair	849 mm	772 mm	-77 mm	4453 mm	1980 mm	-2474 mm
Erie	905 mm	817 mm	-88 mm	803 mm	372 mm	-431 mm
Ontario	930 mm	879 mm	-51 mm	1694 mm	1123 mm	-571 mm
Basin	**Over-Lake Evaporation[a] and Absolute Differences**			**Net Basin Supply[a] and Relative Changes**		
	$1xCO_2$	$2xCO_2$	Diff.	$1xCO_2$	$2xCO_2$	Diff.
Superior	561 mm	736 mm	175 mm	871 mm	684 mm	-21 %
Michigan	647 mm	854 mm	207 mm	813 mm	336 mm	-59 %
Huron	627 mm	829 mm	202 mm	1079 mm	562 mm	-48 %
St. Clair	936 mm	1234 mm	298 mm	4367 mm	1517 mm	-65 %
Erie	898 mm	1197 mm	299 mm	810 mm	-8 mm	-101 %
Ontario	665 mm	874 mm	209 mm	1959 mm	1127 mm	-42 %

[a]Expressed as depths over the lake.

here as the same as over-land, Table 13 shows the same relations for $2xCO_2$ vs. base case precipitation as does Table 8. Net basin supply in Table 13 is less under the $2xCO_2$ climate than under the $1xCO_2$ climate on all lakes. This is true throughout the year for Lake Erie and nearly true on the other lakes (see Figure 15). Net basin supplies are lower throughout the annual cycle except for February and March on Lake Superior, February on Lakes Michigan and Huron, January on Lake St. Clair, and January and February on Lake Ontario. Table 14 summarizes the changes in the hydrologic and net basin supply components for the entire Great Lakes basin; they were computed by converting the equivalent depths of Table 13 to annual flow rates on each lake and adding them over all the lakes. The changes from the base case are also expressed relatively in Table 14. Also expressed relatively are changes from other studies that used other GCMs (*Croley* 1990); they are provided for comparison. Net basin supplies to all Great Lakes are seen to drop to about one half under the CCC GCM; this corresponds to the GCM from the Geophysical Fluid Dynamics Laboratory (GFDL) in the earlier studies. This drop in net basin supply seems to result from the increases in lake evapo-

Table 14. Average Annual Steady-State Great Lakes Basin Hydrology and Net Basin Supply Components.

Scenario	Over Land Precipitation (cms)	Evapotranspiration (cms)	Basin Runoff (cms)	Over Lake Precipitation (cms)	Over Lake Evaporation (cms)	Net Basin Supply (cms)
$1xCO_2$	13815	7825	5987	6604	4992	7783
$2xCO_2$[a]	13598	9518	4077	6578	6587	4180
CCC[a]	-2%	22%	-32%	0%	32%	-46%
GISS[b]	2%	21%	-24%	4%	27%	-37%
GFDL[c]	1%	19%	-23%	0%	44%	-51%
OSU[d]	6%	19%	-11%	6%	26%	-23%

[a]Canadian Climate Centre, prepared by *Louie* (1991).
[b]Goddard Institute for Space Studies GCM, used by *Croley* (1990).
[c]Geophysical Fluid Dynamics Laboratory GCM, used by *Croley* (1990).
[d]Oregon State University GCM, used by *Croley* (1990).

ration and overland evapotranspiration (reducing subsequent runoff to the lakes) observed in the $2xCO_2$ scenario from the GCM.

4.7 Sensitivities

Without temperatures below freezing, the snowpack is insensitive to precipitation. Although the steady-state scenarios on different lakes show different estimates of precipitation change, each shows increases in air temperatures that significantly reduce the snowpack, especially in the southern basins. Thus, even if precipitation increases more than suggested by the GCM, the snowpack will be much reduced under warmer winters. Similarly, regardless of actual changes in precipitation, the Great Lakes basin will experience reduced soil moisture storage and runoff. Both peak shortly after snow melt and then drop throughout the summer and fall due to high evapotranspiration; each climate scenario produces earlier snow melt and a longer period of evapotranspiration. Soil moisture and runoff are most sensitive to precipitation in midsummer when at annual minimums. Thus, within the limits of precipitation produced by the GCMs, soil moisture and runoff scenarios are relatively insensitive to precipitation.

Of the meteorological variables that affect lake evaporation (air temperature, humidity, cloud cover, and wind speed) wind speed is probably the most critical, although air temperature and humidity are also important. Across all lakes and scenarios, daily evaporation was reduced (compared to the base case) only when the scenarios showed reduced wind speeds. If wind speeds remain near historical levels, evaporation will still increase somewhat, however, due simply to the increase in air temperatures that then increase water temperatures. Thus, within the range of other meteorologic variables shown by the GCMs, only if wind speeds are less than historical levels by about 0.5% will lake evaporation not increase.

Because net basin supplies are a sum of lake evaporation, runoff, and precipitation, they are equally sensitive to changes in any of the components. Thus, as long as wind speeds are not much less than historical levels, regardless of precipitation changes (unless precipitation increases are much larger than any shown by the GCMs), net basin supplies are likely to drop due to higher air temperatures that increase evaporation and decrease runoff.

4.8 Summary Comments on Great Lakes Response to Climate Change

The study results should be received with caution as they are, of course, dependent on the GCM outputs, which have large uncertainties. The linkage method used here does not recognize interdependencies between meteorological variables. It also simply changes the magnitude of meteorological time series without affecting their temporal structures. Therefore, changes in variabilities that would take place under a changed climate are not addressed. Seasonal timing differences in the GCM for the changed climate are not reproduced with this method of coupling. Instead, while seasonal meteorology patterns are preserved in the $2xCO_2$ scenario as they exist in the $1xCO_2$ historical data, one still can observe seasonal changes induced by storage effects. Water temperatures increase and peak earlier; heat resident in the deep lakes increases throughout the year. Mixing of the water column diminishes, as most of the lakes become mostly monomictic, and lake evaporation increases. Changes in annual variability are less clear, again as a result of using the same historical time structure for both the base case and the changed climate scenarios.

The higher air temperatures under the $2xCO_2$ scenarios lead to higher over-land evapotranspiration and lower runoff to the lakes with earlier runoff peaks since snowpack is reduced up to 100% and the snow season is shortened more than one month. This also results in a reduction in available soil moisture. Water surface temperatures peak earlier and are higher, with larger amounts of heat resident in the deep lakes throughout the year. Also, buoyancy-driven turnovers of the water column do not occur as often on all lakes except St. Clair. Without biannual turnovers, hypolimnion chemistry may be altered; oxygen may be depleted, releasing nutrients and metals from lake sediments. The lakes may experience more than a single winter turnover if temperature gradients are small and winds are strong enough to induce mixing (*Hutchinson* 1957). Ice formation will be greatly reduced over winter on the deep Great Lakes, and lake evaporation will increase; average steady-state net supplies drop.

5. Coupled Hydrosphere Atmosphere Research Model (CHARM)

The linkage between a GCM and hydrology models allows no feedbacks between these independent models. While the GCMs have crude hydrologic process models, they represent inappropriately large scales and use very simplified conceptualizations. The regional hydrologic impact models may do a much better job of representing the hydrology of an area. However, their use with GCM outputs does not allow the GCM simulations to benefit from these refined processes. Feedback from land and lake surfaces' hydrometeorological properties cannot exist without incorporating regional hydrology models into atmospheric models.

Modelers are turning their attention to mesoscale atmospheric models to enable better assessment of local to regional effects. The leading approach now is to embed mesoscale atmospheric models within GCMs for a region of interest and to couple relevant surface hydrology models to the mesoscale atmospheric model (*Dickinson et al.* 1989; *Giorgi* 1990; *Hostetler et al.* 1993). This allows both the use of more relevant scales for regional impact estimation and the consideration of dynamic linkages between the atmosphere and the surface, now recognized as essential in describing the hydrology and meteorology of an area. This approach has generally been limited in the past to 50-km grids or larger because of the complexity of the modeling system that is required and because of the computer power that was required. The science panel of the GEWEX Continental-Scale International Project and the WMO-CAS Working Group on Numerical Experimentation launched their joint Project for Intercomparison of Land-Surface Parameterization Schemes (PILPS). The National Center for Atmospheric Research (NCAR) is exploring possibilities of operating their atmospheric, hydrologic, and lake flux models embedded in their GCM at scales finer than 50 km.

To estimate impacts associated with both large and fine scales of parameter changes, the Great Lakes research community can address these scales separately. This offers the advantage that we can begin now to look at large-scale parameter changes (such as lake levels, lake-wide heat storage, and annual and monthly water and energy balances) by combining *existing* process models appropriate to these scales. This can be underway while fine-scale parameter changes are investigated. They will require more development and integration of process models. Thus, we have two components to physical modeling of climate-change impacts over the Great Lakes. The first is the integration and use of existing Great Lakes hydrologic process models (lumped-parameter, applying to irregular-shaped areas over spatial scales of 30-100 km for the land surface and 100-300 km for the lake surface and temporal

scales of 1-30 days). The second is the development and integration of fine-scale second-generation (gridded) surface hydrologic process models (at scales from 1-30 km or for simulations of many years) with atmospheric mesoscale models.

5.1 Large-Scale Parameter Changes

We must explore linkages to atmospheric models for existing large-scale irregular-area surface models that already represent excellent portrayals of the hydrology and lake thermodynamics in the Great Lakes. Since hydrological models now exist for large-scale parameter change estimates, large-scale couplings will be useful in beginning derivative studies (such as socio-economic, food-web dynamics, and other secondary impacts identified as dependent on large-scale parameter changes). They also will prove useful as a starting point for subsequent second-generation joint atmospheric-hydrological parameterizations and in the verification of same and of like developments by other investigators. They also will be useful as a base-line for comparing multiple approaches in modeling the atmosphere and hydrology.

GLERL, in cooperation with the Air Resources Laboratory (ARL), is now linking their hydrology models with the Regional Atmospheric Modeling System or "RAMS" (*Pielke et al.* 1990; *Lyons et al.* 1990, 1991a,b). The combination will be used for large-scale parameter investigations, requiring assessment of the temporal and spatial incompatibilities that exist between mesoscale meteorological and regional hydrology models. A modest target is to arrange for coupled modeling by using a 40-km grid, with time steps of 90 seconds in the atmospheric component, coupled to some components of the surface models defined over irregular areas on 12- to 24-hour intervals. RAMS-predicted atmospheric momentum, temperature, moisture, and precipitation fields will be input to the large-scale hydrological models which will use these fields to update sea surface temperature, soil moisture, and snowpack variables. These hydrological parameters will then be input into RAMS to drive the surface energy fluxes over both land and water. Since there is some overlap in function between parts of the atmospheric model and the surface models, decisions are required about which model should be used for some purposes; this is discussed elsewhere (*Croley and Lofgren* 1994).

5.2 Second-Generation Fine-Scale Atmospheric-Hydrologic Integrations

Only when sufficiently fine grids become available for surface hydrology models will surface runoff at points into the lakes be directly estimable from purely gridded models. These fine grids will be approached in the next few years. Likewise, lake heat storage models for the

Great Lakes exist at several levels, from one-dimensional superposition models to three-dimensional circulation models. Again, researchers are approaching fine grids that are usable in long continuous simulations.

Two fine-scale approaches are possible now. The first uses developing atmospheric-hydrologic mesoscale models to estimate joint meteorology and hydrology for surface areas of interest in the Great Lakes and then refines the hydrological estimates through use of the better-calibrated GLERL hydrology models for the Great Lakes. This approach is similar to that taken in linking hydrology models to GCM outputs, described previously. Again, there is no dynamic interaction between the final hydrology models and the atmospheric model. Outputs from the joint atmospheric-hydrologic mesoscale model are inputs to the hydrology models. However, better agreement should be possible since the scales of both sets of models are closer than was true in the GCM-hydrology model studies.

The second fine-scale approach consists of developing second-generation fine-scale Great Lake hydrologic and lake thermodynamic models on finer grids to interface directly with atmospheric models applied at ever-finer resolution and of assessing the importance of two-way runoff-atmospheric interactions unique to CHARM. These will complement similar efforts elsewhere (NCAR) that use alternate models. The matching of spatial and temporal scales between models will proceed at different levels. Linkage will begin with coarse irregular spatial and temporal scales, where existing hydrological models are established over large areas in the Great Lakes (as in the above section), and proceed to finer scales as hydrological models are redeveloped in atmospheric-hydrologic studies. Comparisons will be made between scales to see what is resolved and which process refinements make no difference with regard to different uses (water level estimation, sea breeze predictions, and so forth). Both the atmospheric and hydrological models will be run in three dimensions on the same grid. The grid spacings will be reduced from 30 km to 15, 10, 5, and 1 km scales. For the smaller scales, non-hydrostatic physics and explicit cloud microphysics will be employed. To start out, interactions will be performed at the time step of the atmospheric model (between 5-90 seconds depending on the horizontal resolution of the grids). Sensitivity experiments will be performed to determine an optimum update frequency between the atmospheric and hydrologic models since it may not be necessary to interact the models every time step.

5.3 Summary Comments on CHARM

Earlier assessments used atmospheric GCM outputs as meteorologic scenarios to drive process models for generating hydrologic scenarios. Climate change effects were inferred by comparing process model outputs for a base case with the changed climate scenario. As the

linkage methods of these assessments constrained spatial and temporal meteorologic variabilities to those present in the historical records, impact assessments began with the transference of existing climates to the Great Lakes. Lack of feedback between surface process models and atmospheric models is still a problem.

Researchers are now developing and verifying multi-scale hydrologic models, with appropriate links to mesoscale atmospheric models, using spatially extensive observations based upon satellite and *in situ* measurements and supported by field experiments. These linked models are slated to be embedded in GCM or other boundary condition simulations to assess climate change effects. GLERL is working with ARL to investigate alternative CHARM possibilities. Now underway are a large-scale coupling, that employs GLERL's existing irregular-area surface models, and a series of finer-scale couplings where surface models are defined over the same (surface) grid as used in the mesoscale atmospheric models. GLERL also plans to work with existing and developing coupled atmospheric-hydrologic mesoscale models over the Great Lakes by refining hydrologic estimates with more-detailed hydrologic and lake surface flux models.

6. Acknowledgments

This paper is GLERL contribution no. 884.

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MODELING OF RUNOFF AND STREAMFLOW AT REGIONAL TO GLOBAL SCALES

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ABSTRACT

Global change problems place a new set of demands on hydrologic models. The primary purpose for representation of land surface hydrology in the context of coupled land-atmosphere models is to partition downward solar and longwave radiation into latent, sensible, and ground heat fluxes, and upward longwave radiation, rather than to predict streamflow. Nonetheless, past work in the development and application of conceptual streamflow simulation models for operational applications, such as forecasting, can provide valuable lessons, especially with respect to model parameter parsimony, for the development and application of land surface parameterizations for coupled land-atmosphere models. Some important issues in model development and application are illustrated in the context of the Variable Infiltration Capacity two-layer (VIC-2L) model. Application of the VIC-2L model to FIFE (central Kansas Grassland) and ABRACOS (cleared Amazonia tropical forest) field data are described. A version of VIC-2L that incorporates streamflow routing is described, along with some results of its application off-line (climatalogical forcing) to the Columbia River basin. Finally, an approach for regional estimation of the parameters of VIC-2L is described, along with preliminary results for an application to the Columbia River basin (drainage area approximately 615,000 km^2) at a one degree by one degree spatial resolution.

1.0 INTRODUCTION

This paper focuses on dynamic hydrologic modeling at regional to global scales. By dynamic I mean time scales of not more than a day, and by regional I mean spatial domains that include at least several grid cells of a numerical weather prediction or climate model, that is, length scales of at least several hundred km. Both the threshold time and space scales are recognized to be arbitrary. The paper consists of four parts: 1) a review of the relevance of past work in operational hydrology in the context of streamflow simulation at regional to global scales; 2) some key issues and approaches to land surface parameterizations for large scale atmospheric models; 3) an outline of an approach to modeling large North American rivers with preliminary results for the Colorado River; and 4) some ideas and early results from the use of regionalization methods to estimate the parameters of a coupled land-atmosphere model.

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Preface

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